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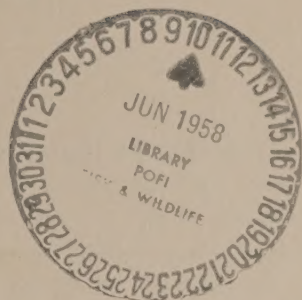
神戸海洋气象台

歐 文 報 告

第 1 2 卷 第 2 号

THE
MEMOIRS

OF THE
KOBE MARINE OBSERVATORY
KOBE, JAPAN



Vol. XII No. 2

K O B E

March. - 1958

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On the Fluctuation of the Kuroshio South of Honshū (2)

(Seasonal variations of water temperature and
chlorinity in the upper layer)

SHIGEO MORIYASU

1. Introduction

Many, useful studies have been made about the annual variation in the oceanographic elements of the Kuroshio in the considered sea area and the outline has been made clear. Koenuma had a series of works regarding the Kuroshio, in which seasonal variation of water temperature was considered (1939). Koizumi (1955) treated oceanographic data of the ocean weather station "Tango" (29°N , 135°E) and discussed characteristics of the annual variation in the oceanographic elements in detail. Fukuoka and Tsuiki (1954), and Ichiye (1956) discussed the annual variation of chlorinity in their interesting way.

We will study the seasonal variation of water temperature and chlorinity in the upper layer mainly. Here, the dynamical consideration is not made but the diffusion equation is used to explain the variation.

2. Distributions of water temperature and chlorinity

Distributions of water temperature, chlorinity and density *in situ* should be discussed by using the results of oceanographic observations in each season. Here, the data in August and November, 1956, and in March and May, 1957, are used (J. M. A.). The oceanographic conditions in these periods have been reported (J. M. A. 1956~7). The

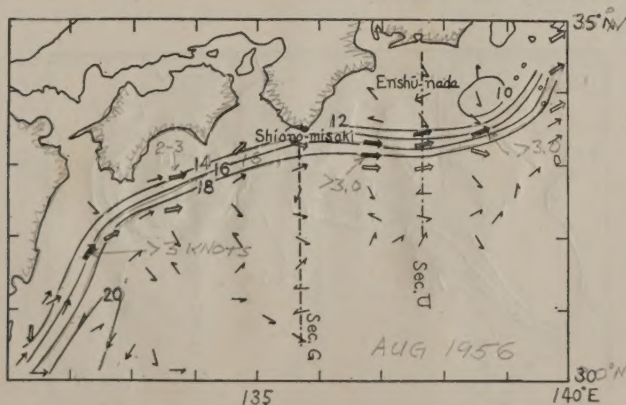


Fig. 1. Water temperature in the 200m layer and surface current with GEK in August, 1956. Full line is isothermal line and arrow indicates surface current as follows:

$\leftarrow < 0.9\text{kt}$, $\leftarrow 1.0 \sim 1.9\text{kt}$, $\leftarrow 2.0 \sim 2.9\text{kt}$, $\leftarrow > 3.0\text{kt}$.

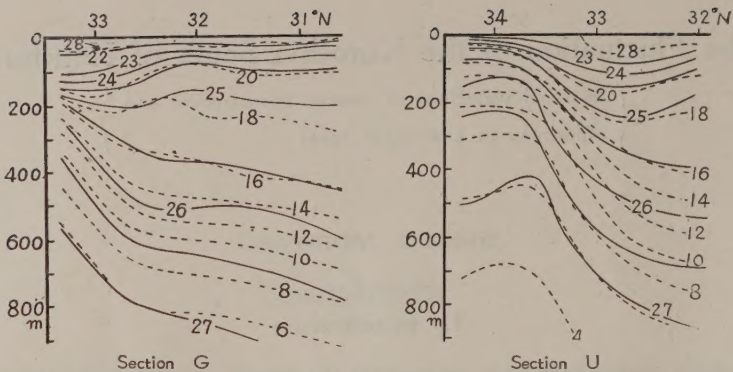


Fig. 2. Vertical sections of water temperature (broken line) and density *in situ* (Full line) in August, 1956.

characteristic features in each season are made clear as follows.

August, 1956 (Figs. 1 and 2). There was a cold upwelling region in Enshū-nada, where a counter-clockwise eddy was found. The Kuroshio was near the coast west of Shionomisaki and ran eastwards without meandering in the offing of Enshū-nada. In the coastal and upwelling region, the thickness of isothermal layer was very small and at most 20~30 m, and beneath this layer there was transitional layer where water temperature decreased downwards. On the other hand, in the off-shore warm region, a layer with a little larger vertical gradient existed between the isothermal and transitional layer noted above, and the isothermal layer was thicker than that in the coastal and upwelling region. These vertical distributions were common to those of density, also. In the coastal and upwelling region, the thin homogeneous layer was found above the transitional layer with a large vertical stability, while, in the off-shore warm region, a large vertical stability was found between homogeneous and transitional layer. Near the surface, the thin layer with extremely small density under the influence of decreasing in chlorinity was observed.

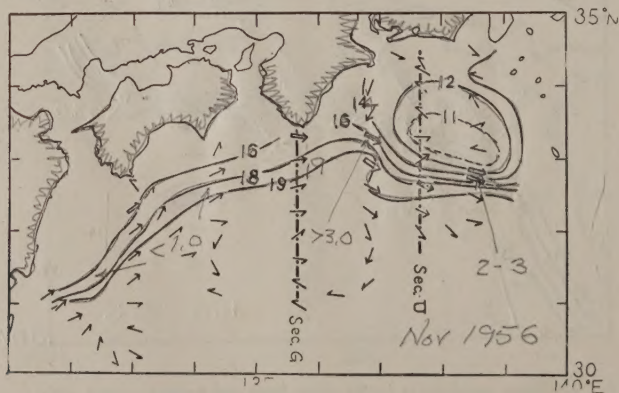


Fig. 3. Water temperature in the 200m layer and surface current with GEK in November, 1956. Notation is the same as in Fig. 1.

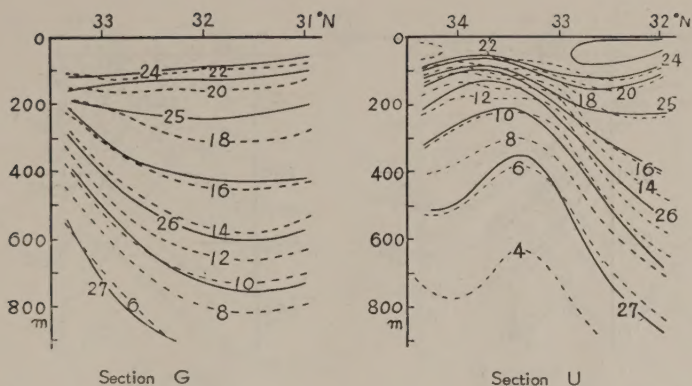


Fig. 4. Vertical sections of water temperature (broken line) and density *in situ* (full line) in November, 1956.

November, 1956 (Figs. 3 and 4). The cold region with a counter-clockwise eddy was found distinctly in Enshū-nada. The current axis existed near the coast west of Shionomisaki, while it shifted southward a little off Enshū-nada. In general, the homogeneous or isothermal layer near the surface became thicker than in August. Hence, the depth in which the large vertical gradient and large stability were found was larger. In this season, the homogeneous layer was growing and its thickness was about 100 m. But, the other conditions were almost equal.

March, 1957 (Figs. 5 and 6). A weak counter-clockwise eddy was found in Enshū-nada where water temperature was lower than that in the other regions. The Kuroshio flowed extremely near the coast off Shionomisaki and the meandering was not found off Enshū-nada. The most interesting characteristic feature was the growth of homogeneous layer. In this season, the homogeneous layer was very thick over the sea area. Namely, the thickness was about 100 m in the coastal and upwelling region, while it was larger than 400 m in the off-shore warm region. Vertical distributions were the most simple in the year, that is, the homogeneous layer was in contact with transitional

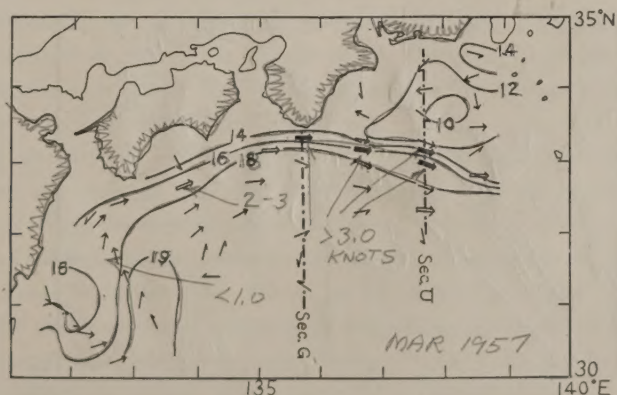


Fig. 5. Water temperature in the 200m layer and surface current with GEK in March, 1957. Notation is the same as in Fig. 1.

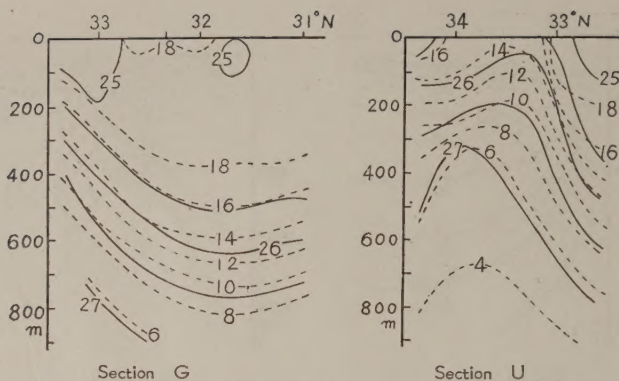


Fig. 6. Vertical sections of water temperature (broken line) and density *in situ* (full line) in March, 1957.

layer. In the upper layer, the large vertical stability vanished and the completely homogeneous layer was produced, and then the distributions were quite different from those in summer. In the transitional layer, however, the striking variation did not occur except for one relating to dynamical effect.

May, 1957 (Figs. 7 and 8). Though the oceanographic conditions were almost the same as those in March, the counter-clockwise eddy in the cold region was indistinct. A short period fluctuation was found clearly near Shiono-misaki, that is, the westerly current of 0.9kt was transformed into the easterly current of 3.7kt at $33^{\circ}20' \text{ N}$, $135^{\circ}40' \text{ E}$ and the current axis shifted northward during about a week. The homogeneous layer could not be found in strict sense, but the vertical stability near the surface was relatively small. The transitional layer lay adjacent to the homogeneous layer and the stable stratification began to be formed in the coastal and upwelling region, while another layer with larger vertical stability was found in the depth of about 100 m between both layers in the off-shore warm region.

The variation in the distribution of water temperature and density mentioned above

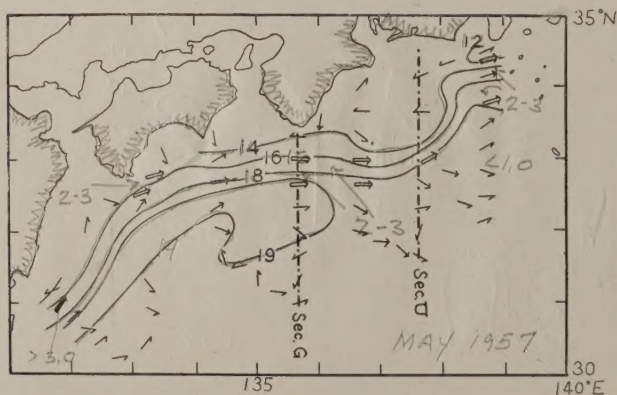


Fig. 7. Water temperature in the 200m layer and surface current with GEK in May, 1957. Notation is the same as in Fig. 1.

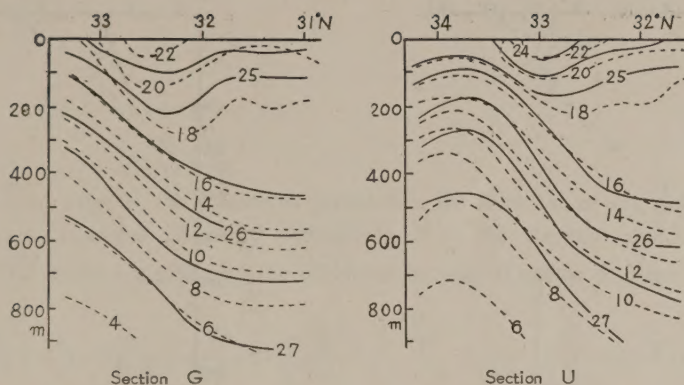
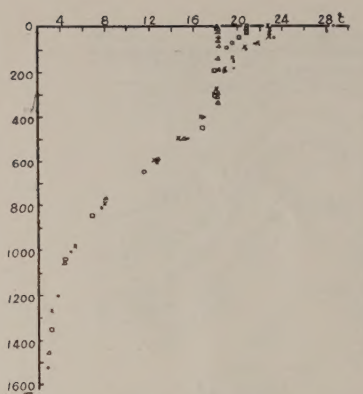


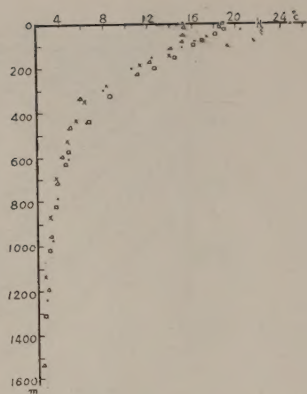
Fig. 8. Vertical sections of water temperature (broken line) and density *in situ* (full line) in May, 1957.

is summarized as follows. In winter, the homogeneous layer develops most strongly, in which temperature and chlorinity are nearly constant, so that density *in situ* is also constant. In spring, the water temperature increases on account of the heating at the sea surface and the homogeneous layer is destroyed near the surface. However, as the effect does not extend to the whole layer, the homogeneous layer is maintained in the lower part of the layer in the off-shore warm region where the homogeneous layer is very thick in winter. From spring to summer, the tendency is strengthened more and more, and the vertical stability increases near the surface owing to the heating and the decrease of chlorinity. The phenomenon is especially striking in the coastal and upwelling region. When the cooling begins at the surface from summer to autumn, the water temperature near the surface decreases and the density becomes larger than that in the lower layer. So, an unstable condition is formed and convection is produced, and then the homogeneous layer grows rapidly. By these processes, a completely homogeneous layer is formed till the depth of about 100 m in the coastal and upwelling region. In the off-shore region, however, the same conditions as those in summer are kept below the depth, because the effect of convection does not extend to sufficient depth. From autumn to winter, the process continues and the thickness of the homogeneous layer grows, and finally the homogeneous layer is formed completely to the depth of about 400 m in the off-shore region, too.

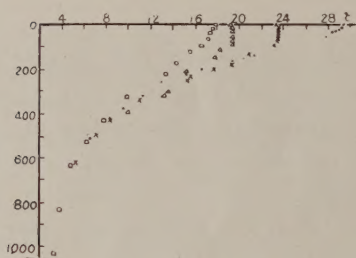
Accordingly, the thermal influence from the surface is limited in the layer noted above and does not reach to the layer below the depth. In other words, its thickness is the depth to which the thermal effect can reach. It is about 100 m and 400 m in the coastal and off-shore region, respectively. Though it depends on the intensity of upwelling, it seems to be less than 100 m in the upwelling region for the example noted here. The values of water temperature, chlorinity and density *in situ* in the depth are estimated nearly from vertical sections as follows: $t = 16^{\circ}\text{C}$, $Cl = 19.20\text{‰}$ and $\sigma_t = 25.5$. These values seem to be critical values except for especial examples. Therefore, the layer with $\sigma_t > 25.5$ can be taken as the transitional layer where the external and thermal effect do not appear and dynamical effect predominates. This will be ascertained



off-shore warm region (31°00'N, 135°40'E)



upwelling region (around 34°00'N, 137°40'E)



coastal region (33°20'N, 135°40'E)

Fig. 9. Vertical distributions of water temperature at 3 oceanographic stations.

(• August, 1956; × November, 1956; △ March, 1957; □ May, 1957)

by T-CI relations in section 5.

Three examples of vertical distribution of water temperature are shown in Fig. 9, which are typical ones in the coastal, upwelling and off-shore warm region, respectively. As the vertical distributions are closely related to the position of the current axis of the Kuroshio and that of the cold region, their fluctuations affect the vertical distributions of water temperature. In the off-shore warm region noted here, the dynamical effect hardly occur and it seems that only the thermal effect is predominant, and then water temperature in the depth below 400 m does not indicate seasonal variation. For other two examples, however, the dynamical effects (for example, the movement of the upwelling region and the fluctuation in the intensity of upwelling, etc.) appear to a certain extent in the considered period and we find comparatively striking fluctuations of water temperature in the transitional layer.

3. Rise of water temperature from winter to summer

From winter to summer, the stable upper layer is heated from the surface and the effect extends downwards gradually by eddy diffusion. The rise of water temperature in the layer should be considered by using a simple model. If a constant heat flux is given at the sea surface and horizontal diffusion and advection are not considered, we have the following equation relating to the variation of water temperature θ ;

$$\frac{\partial \theta}{\partial t} = k \frac{\partial^2 \theta}{\partial z^2}$$

with

$$\left\{ \begin{array}{ll} \frac{\partial \theta}{\partial z} = -\frac{f_0}{K} & \text{at } z=0 \\ \theta = f(z) & \text{when } t=0, \end{array} \right.$$

where k is mean value in the considered period, f_0 is the heat flux at sea surface ($z=0$), z is the depth measured positively downwards, K is the thermal conductivity and $k = \frac{K}{\rho}$. The solution is given by the following formula (Carslaw & Jaeger, 1947):

$$\theta = -\frac{f_0}{K} \left[z \left\{ 1 - \Phi \left(\frac{z}{2\sqrt{kt}} \right) \right\} - \frac{2\sqrt{kt}}{\sqrt{\pi}} e^{-\frac{z^2}{4kt}} + \frac{1}{\sqrt{\pi}} \int_{\frac{z}{2\sqrt{kt}}}^{\infty} f\left(z + 2\sqrt{kt}\lambda\right) e^{-\lambda^2} d\lambda \right. \\ \left. + \frac{1}{\sqrt{\pi}} \int_{\frac{z}{2\sqrt{kt}}}^{\infty} f\left(-z + 2\sqrt{kt}\lambda\right) e^{-\lambda^2} d\lambda \right],$$

where $\Phi(x) = \frac{2}{\sqrt{\pi}} \int_0^x e^{-t^2} dt$ (error function). The first term in the right hand side indicates the effect of the heat flux at the surface and the other terms relate to the initial distribution. In winter, the homogeneous layer is predominant and the water temperature is nearly homogeneous from surface to a certain depth. So the initial condition is assumed as follows:

$$f(z) = \begin{cases} \theta_0 \text{ (constant)} & \text{at } 0 \leq z \leq z_0 \\ \theta_0 - \beta(z - z_0) & \text{at } z \geq z_0, \end{cases}$$

where z_0 is the depth of homogeneous layer. On such conditions, the solution leads to

$$\theta - \theta_0 = -\frac{f_0}{K} \left[z \left\{ 1 - \Phi \left(\frac{z}{2\sqrt{kt}} \right) \right\} - \frac{\sqrt{kt}}{2} \Phi_1 \left(\frac{z}{2\sqrt{kt}} \right) - \frac{\beta}{2} \Phi \left(\frac{z - z_0}{2\sqrt{kt}} \right) \cdot (z - z_0) \right. \\ \left. - \frac{\beta}{2} \Phi \left(\frac{z + z_0}{2\sqrt{kt}} \right) \cdot (z + z_0) + \beta z_0 - \frac{\beta \sqrt{kt}}{2} \left[\Phi_1 \left(\frac{z - z_0}{2\sqrt{kt}} \right) + \Phi_1 \left(\frac{z + z_0}{2\sqrt{kt}} \right) \right] \right],$$

where Φ_1 is the first derivative of the error integral, that is, $\Phi_1(x) = \frac{2}{\sqrt{\pi}} e^{-x^2}$. As the rise of water temperature at the sea surface ($z=0$) is $\Delta\theta_0 = \theta_{z=0} - \theta_0$,

$$\frac{f_0}{K} = \frac{\sqrt{\pi}}{2\sqrt{kt}} \left[\Delta\theta_0 + \beta z_0 \left\{ \Phi \left(\frac{z_0}{2\sqrt{kt}} \right) - 1 \right\} + \beta \sqrt{kt} \Phi_1 \left(\frac{z_0}{2\sqrt{kt}} \right) \right].$$

Therefore, the rise of water temperature at $z = nz_0$ ($0 < n \leq 1$).

$\Delta\theta_z = \theta_{z=nz_0} - \theta_0$, is represented by

$$\frac{\Delta\theta}{\beta z_0} = -\sqrt{\pi} \left[n\alpha_0 \left\{ 1 - \Phi(n\alpha_0) \right\} - \frac{1}{2} \Phi_1(n\alpha_0) \right] \left[\frac{\Delta\theta_0}{\beta z_0} + \Phi(\alpha_0) + \frac{1}{2\alpha_0} \Phi_1(\alpha_0) - 1 \right] \\ + 1 - \frac{1-n}{2} \Phi\{(1-n)\alpha_0\} - \frac{1+n}{2} \Phi\{(1+n)\alpha_0\} - \frac{1}{4\alpha_0} \left[\Phi_1\{(1-n)\alpha_0\} + \Phi_1\{(1+n)\alpha_0\} \right],$$

where $\alpha_0 = \frac{z_0}{2\sqrt{kt}}$. The solution is the function of α_0 only, because $\Delta\theta_z$, $\Delta\theta_0$, β and z_0 are obtained from observed data. Hence α_0 can be easily found then k and f_0

are computed. Here, $\rho \doteq 1$.

The introduced results are applied to station A in the warm region south of Shionomisaki and station B in the cold region off Enshū-nada, where dynamical effects on the distribution of water temperature may be neglected in the considered period. The

Station	Period	Initial condition		Computed values			
		z_0	β	α_0	k	f_0	$F_0 = f_0 t$
A	March 14~May 9	3.5×10^4	2.12×10^{-4}	2.0	15	4.3×10^{-8}	2.11×10^4
	March 14~August 1	"	"	2.7	3.6	5.7	6.84
	May 9~August 1	—	—	—	—	6.7	4.73
B	March 21~August 6	0.8	3.65	1.2	1.0	2.8	3.32

Table 1. Computed values of eddy conductivity and heat flux in the warm region (station A) and in the cold water region (station B). Here, the units are C. G. S. and $^{\circ}\text{C}$.

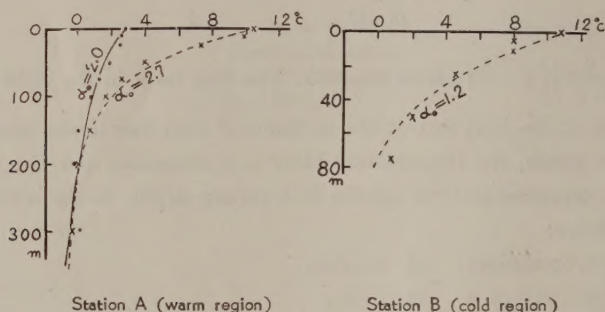


Fig. 10. Comparison between observed and computed distribution of water temperature.

observed value: $\begin{cases} \bullet & \text{May, 1957} \\ \times & \text{August, 1957} \end{cases}$
 computed curve: $\begin{cases} \text{—} & \text{May, 1957} \\ \text{---} & \text{August, 1957} \end{cases}$

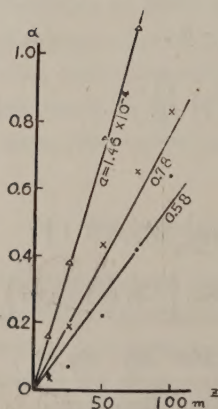


Fig. 11. Relation between α and z when heating at the sea surface is considered only. Full lines are obtained by method of the least squares on the assumption that

$$\alpha = \frac{z}{2\sqrt{kt}} = \alpha z.$$

- Values of α obtained from March to May, 1957 at station A
- $\times$ Values of α obtained from March to August, 1957 at station A
- $\triangle$ Values of α obtained from March to August, 1957 at station B

obtained values are given in Table 1 and the comparison between observed and computed distribution of water temperature is shown in Fig. 10.

The results mentioned above are obtained by solving the diffusion equation exactly. But, other terms except for that relating to the heat flux at the surface in the solution seem to be effective only in the lower part. So, if the initial condition is $\theta = 0$, the following relation is introduced:

$$\frac{\Delta\theta_z}{\Delta\theta_0} = e^{-\alpha^2} - \sqrt{\pi} \alpha \left\{ 1 - \Phi(\alpha) \right\}$$

where $\alpha = \frac{z}{2\sqrt{kt}}$ and $\Delta\theta_0 = \frac{2f_0}{K} \left(\frac{kt}{\pi} \right)^{\frac{1}{2}} \cdot \frac{\Delta\theta_z}{\Delta\theta_0}$ and then α can be obtained from observations. The relationship between α and z for the above examples is shown in Fig. 11, which indicates that they correlate linearly with each other. Accordingly, kt can be computed by the method of least squares on the assumption that $\alpha = az$ is admitted. The values of k obtained in this case are nearly the same as those in the previous paragraph.

Now, we will estimate the error of the eddy conductivity and the heat flux at the sea surface. We have

$$\frac{\Delta\alpha_0}{\alpha_0} = \frac{\Delta z_0}{z_0} - \frac{\Delta k}{2k} - \frac{\Delta t}{2t}$$

from $\alpha_0 = \frac{z_0}{2\sqrt{kt}}$ by logarithmic differentiation. As $\frac{\Delta z_0}{z_0}$ and $\frac{\Delta t}{t}$ are negligible as compared with $\frac{\Delta\alpha_0}{\alpha_0}$,

$$-\frac{\Delta k}{k} = -2 \frac{\Delta\alpha_0}{\alpha_0}$$

is led. In the same way, the following relation as to f_0 is obtained:

$$-\frac{\Delta f_0}{f_0} = -\frac{\Delta k}{2k} = \frac{\Delta\alpha_0}{\alpha_0}$$

Here, $f_0 = \frac{\rho_1 \sqrt{\pi k}}{2\sqrt{t}} \Delta\theta_0$ is used. The error of α_0 seems to be ± 0.1 . At station A,

$\frac{\Delta k}{k} = 0.1$ is obtained because of $\frac{\Delta\alpha_0}{\alpha_0} = 0.05$ and the error of k is expected to be

about 1.5. While at station B, $\frac{\Delta k}{k} = 0.2$ and k may have the error of about 0.2.

Similarly, the error of f_0 is found to be $2 \sim 3 \cdot 10^{-4}$ at both stations. Taking into account of the error of k and f_0 , we will glance at their computed values. At station A, k is larger from March to May than from March to August and this agrees with the decrease of k owing to the increase of vertical stability. Furthermore, k in the cold water region with large vertical stability is smaller than that in the warm region. On the other hand, the heat flux f_0 at station A from March to May is about two thirds of that from May to August. From March to August, the heat flux at station A is nearly twice that at station B. Here, the total flux $F_0 = f_0 t$ from May to August as computed as the difference between that from March to August and that from March to May and then the mean heat flux f_0 is obtained.

In the above consideration, we prepare unsuitable assumptions such as the constant

heat flux at the sea surface and constant eddy conductivity. Nevertheless, the obtained results do not contradict the observed fact remarkably. So, the rise of water temperature in the upper layer may be considered to depend upon the process stated above.

4. Growth of homogeneous layer from summer to winter

From summer to winter, the surface water temperature falls owing to the cooling at the sea surface and vertical stability decreases because of the increase in density and finally an unstable condition is formed. Hence, the vertical mixing develops and the homogeneous layer is produced. The thickness of the layer grows thicker and thicker gradually, while the cooling at the sea surface is kept. Okubo(1953) discussed the process of convection in the lake. In the fresh-water lake, the density is almost dependent upon water temperature only, while in the sea it is related to chlorinity in addition to water temperature. Now, we will estimate the effect of chlorinity on the growth of homogeneous layer on the same assumptions as Okubo's.

At initial stage ($t=t$), h is the thickness of homogeneous layer with temperature θ' , density ρ' , and chlorinity S' . Below the layer, density, temperature and chlorinity is ρ , θ and S , respectively. If the thickness becomes $h + \Delta h$ at $t=t + \Delta t$, we have

$$\rho'_{t=t+\Delta t} + \left(\frac{\partial \rho'}{\partial t} \right)_{t=t} \Delta t = \rho_{z=h} + \left(\frac{\partial \rho}{\partial z} \right)_{z=h} \Delta h.$$

$$\text{Since } \rho'_{t=t+\Delta t} = \rho_{z=h}, \quad \frac{\partial \rho'}{\partial t} = \frac{\partial \rho'}{\partial \theta'} \frac{\partial \theta'}{\partial t} + \frac{\partial \rho'}{\partial S'} \frac{\partial S'}{\partial t} \quad \text{and}$$

$$\frac{\partial \rho}{\partial z} = \frac{\partial \rho}{\partial \theta} \frac{\partial \theta}{\partial z} + \frac{\partial \rho}{\partial S} \frac{\partial S}{\partial z},$$

the equation leads to

$$\left(\frac{\partial \rho'}{\partial \theta'} \frac{\partial \theta'}{\partial t} + \frac{\partial \rho'}{\partial S'} \frac{\partial S'}{\partial t} \right)_{t=t} \Delta t = \left(\frac{\partial \rho}{\partial \theta} \frac{\partial \theta}{\partial z} + \frac{\partial \rho}{\partial S} \frac{\partial S}{\partial z} \right)_{z=h} \Delta h.$$

The orders of the terms in the left hand side are as follows:

$$\frac{\partial \rho'}{\partial \theta'} \frac{\partial \theta'}{\partial t} \Delta t \sim 10^{-3} \quad \text{if } \Delta \theta' \sim 10 \quad \text{and} \quad \frac{\partial \rho'}{\partial \theta'} \sim 10^{-4};$$

$$\frac{\partial \rho'}{\partial S'} \frac{\partial S'}{\partial t} \Delta t \sim 10^{-7} \quad \text{if } \Delta S' \sim 10^{-4} \quad \text{and} \quad \frac{\partial \rho'}{\partial S'} \sim 10^{-4}.$$

Here, the values of $\frac{\partial \rho'}{\partial \theta'}$ and $\frac{\partial \rho'}{\partial S'}$ by Munk and Anderson (1948) are used. Thus, the second term in this side can be neglected and we have

$$\frac{\partial \theta'}{\partial t} \Delta t = \left(\frac{\partial \theta}{\partial z} + \frac{\partial \rho / \partial S}{\partial \rho / \partial \theta} \frac{\partial S}{\partial z} \right)_{z=h} \Delta h,$$

where $\frac{\partial \rho'}{\partial \theta'} = \frac{\partial \rho}{\partial \theta}$ is assumed. Furthermore, the order of the terms in the right hand

side should be estimated. The values of $\frac{\partial \rho / \partial S}{\partial \rho / \partial \theta}$ calculated from the values of $\frac{\partial \rho}{\partial S}$

and $\frac{\partial \rho}{\partial \theta}$ by Munk and Anderson are shown in Table 2. Accordingly, $\frac{\partial \rho / \partial S}{\partial \rho / \partial \theta} \sim 3$

is used in this case. Since $\frac{\partial S}{\partial z} \sim 10^{-8}$ and $\frac{\partial \theta}{\partial z} \sim 10^{-4}$, $\frac{\partial \rho / \partial S}{\partial \rho / \partial \theta} \frac{\partial S}{\partial z} \sim 10^{-8}$ and the

second term is negligible as compared with the first term. So we have

$$\frac{\partial \theta'}{\partial t} \Big|_{t=t} \Delta t = \frac{\partial \theta}{\partial z} \Big|_{z=h} \Delta h$$

and when $\Delta t \rightarrow 0$, the differential equation

$$\frac{\partial \theta'}{\partial t} \Big|_{t=t} = - \frac{\partial \theta}{\partial z} \Big|_{z=h} \frac{\partial h}{\partial t}$$

$\theta \backslash S$	18.26	19.38	20.49‰
0°C	-17.9	-14.6	-13.5
10	-4.58	-4.53	-4.46
20	-2.91	-2.90	-2.88

Table 2. The values of $\frac{\partial \rho / \partial S}{\partial \rho / \partial \theta}$ computed from $\frac{\partial \rho}{\partial S}$ and $\frac{\partial \rho}{\partial \theta}$ by Munk and Anderson.

is introduced. From this, the results in a fresh-water lake obtained by Okubo are also applied to the ocean and the influence of chlorinity need not be considered.

The equation shows that the more the vertical gradient of water temperature is large, the more the rate of the growth in the thickness of homogeneous layer is small. In the cold water region where the vertical gradient of water temperature is large, the growth of homogeneous layer is not so striking as in the warm region with the small vertical gradient. Thus, the thickness of the layer is larger in the warm region than in the cold water region from autumn to winter, because $\frac{\partial \theta}{\partial t}$ is nearly equal in both regions.

5. T-Cl relations.

Seasonal variations of water temperature and density have been stated and their characteristics have been made clear. Next, noteworthy features of T-Cl relations in each season should be discussed on the basis of the same data as in the preceding section.

August, 1956 (Fig. 12). Near the sea surface, water temperature was very high and reached nearly 30°C owing to the solar radiation. On the other hand, the values of chlorinity were within 18.50~19.35‰. Thus the horizontal difference of water temperature was small, while the difference of chlorinity was large as chlorinity decreased strikingly in the coastal region and had almost the same value as that in other seasons in the off-shore region. The difference of chlorinity corresponding to a water temperature decreased with depth, accordingly with decrease of water temperature, and was nearly constant (about 0.05‰) relating to water temperature below 17°C. Maximum of chlorinity was about 19.35‰ and its minimum was about 18.90‰, and the water temperature corresponding to the minimum was about 7°C. The depth where minimum of chlorinity appeared was about 400m in the cold water region while it was below 800m in the off-shore region.

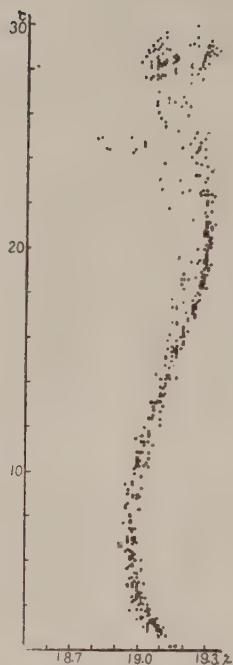


Fig. 12. T-Cl relations in August, 1956.

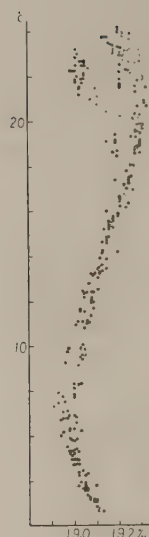


Fig. 13. T-Cl relations in November, 1956.

November, 1956 (Fig. 13). Water temperature was below 24°C and chlorinity near the sea surface was higher than that in August. Minimum of chlorinity near the sea surface was above 19.00‰ , while maximum was about 19.35‰ as in August. The range of the value of the chlorinity corresponding to a water temperature was almost the same as in August, but the large range corresponding to high water temperature was not found. In the layer with water temperature below 18°C , the breadth of the T-Cl curve was nearly constant and minimum of chlorinity was about 18.90‰ which corresponded to water temperature of about 7°C . The depth where the minimum appeared was from 300m to 800m.

March, 1951 (Fig. 14). Water temperature was below 20°C . In the subsurface layer, the range of the value of chlorinity corresponding to a water temperature was small (below 0.1‰) and maximum of chlorinity was 19.30‰ and the value below 19.00‰ was not found. All data were included within a curve with the same breadth as that in August and November, 1956. Minimum of chlorinity was 18.90‰ and appeared in the depth from 300m to 800m where water temperature was $7 \sim 8^{\circ}\text{C}$. The points surrounded by dotted line in Fig. 14 were obtained near the sea surface in the cold water region and were not included within the curve. This is due to the fall of water temperature by convection in the upwelling region.

May, 1957 (Fig. 15). Near the sea surface, we found the rise of water temperature and the fall of chlorinity, and the range of the chlorinity was larger than that in March, but maximum of chlorinity was nearly equal to that in March. In the layer with tem-

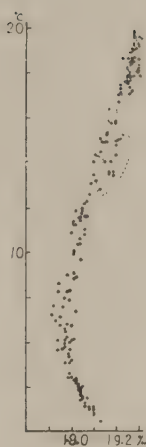


Fig. 14. T-Cl relations in March, 1957

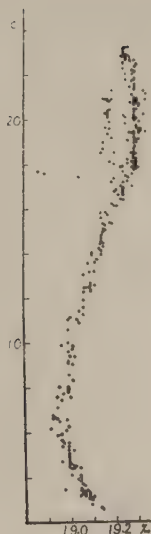


Fig. 15. T-Cl relations in May, 1957

perature below 17°C , the curve was not different from those in the preceding three examples, and minimum of chlorinity was 18.90‰ and appeared in the depth from 400m to 800m where water temperature was about 7°C . From the above, the transitional condition from winter to summer was seen in T-Cl relations. At $33^{\circ}20'\text{N}$, $135^{\circ}40'\text{E}$, water temperature and chlorinity at the surface was below 18°C and below 18.90‰ , respectively. This seemed to relate to dynamical fluctuation in addition to extenal influence.

The T-Cl relations stated above are summarized as follows.

In the layer with water temperature above 16°C .

The range of the chlorinity corresponding to a water temperature is not constant in each season and its maximum and minimum is seen in summer and winter, respectively. Maximum value of chlorinity is nearly 19.35‰ throughout the year and is found in this layer, while its minimum value is found at the surface and depends upon meteorological conditions.

In the layer with water temperature below 16°C .

T-Cl curve does not fluctuate remarkably and a constant relation seems to exist between water temperature and chlorinity. Namely, the value of chlorinity can be determined uniquely as a function of water temperature. The breadth of T-Cl curve is less than 0.1‰ . Minimum of chlorinity is about 18.90‰ and corresponds to the water temperature of about 7°C . We find minimum of chlorinity in the depth from 300m to 800m which is larger in the off-shore warm region than in the coastal and upwelling region. From the mean values of chlorinity within the each interval of 1°C in each season shown in Table 3, it seems that a nearly constant relation is maybe found in the layer. Further, mean values of their mean values within the interval of 1°C are computed. Accordingly, the curve connecting these points shows the mean relation between

t°C	16~15	15~14	14~13	13~12	12~11	11~10	10~9	9~8	8~7	7~6	6~5	5~4	4~3	3~2
August, 1956	19.160	19.134	19.101	19.080	19.035	19.015	18.987	18.980	18.971	18.972	18.990	19.004	19.054	19.111
November, 1956	.171	.150	.118	.070	.058	.040	19.019	19.013	.957	.980	.994	.017	.064	.110
March, 1957	.171	.140	.108	.070	.036	.035	.020	18.980	.960	.973	.979	.018	.053	.101
May, 1957	.160	.138	.100	.083	.034	.003	18.998	.985	.971	.947	.979	.000	.054	.113
mean	19.165	19.141	19.107	19.076	19.041	19.023	19.006	18.989	18.965	18.968	18.970	19.010	19.056	19.109

Table. 3. Mean values of chlorinity(‰) within the each interval of water temperature of 1°C.

water temperature and chlorinity and almost all data are included between the both curves deviated by about $\pm 0.05\%$ from the mean curve.

From the above expression, it is appropriate to distinguish the layer with water temperature above 16°C from the layer with water temperature below 16°C, and the critical point has water temperature of 16°C and chlorinity of 19.20‰, so density *in situ* of 25.50. External influences such as the heating and cooling at the surface and the decrease of chlorinity near the surface can reach to the critical point, which is located in the larger depth in the off-shore, warm region than in the coastal and upwelling region. These agree with the results from the distributions of water temperature and density.

In the layer with water temperature below 16°C, the effect from the exterior can be

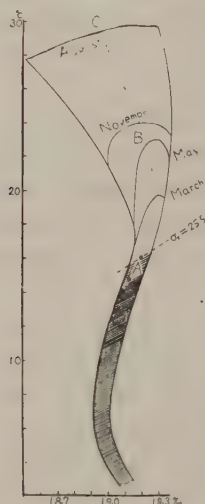


Fig. 16. Schematic representation of T-Cl relations in each season. Shaded portion corresponds to the relations in the transitional layer. Portion surrounded with thin line represents the relations in the upper layer in each season.

hardly found and the dynamical fluctuation has close relation to the depth where the same T-Cl relations appear. On the contrary, a remarkable annual fluctuation of water temperature and chlorinity is seen in the upper layer with water temperature above 16°C. The fluctuation in the upper layer is represented on T-Cl diagram schematically (Fig. 16). From this figure, the following is deduced: In winter, T-Cl relation can be indicated approximately by one point A, since water temperature and chlorinity are nearly homogeneous under the influence of predominant vertical mixing and their horizontal gradient is not large. From winter to summer, water temperature rises and chlorinity falls in general near the surface, and these phenomena extend downwards gradually. This process is kept until summer, but no variation can be found at the critical point. Hence, T-Cl curve at each station is represented

by the curve ABC, where C and B indicates the value at the surface in summer and in spring, respectively. As the amount of the fall of chlorinity is much larger near the coast than in the off-shore region, the curves in the left portion indicate the T-Cl relations near the coast. From summer to winter, a converse process, that is, the fall of temperature and the rise of chlorinity progresses and the curves are reduced finally to a point A in winter. Accordingly, the distribution of chlorinity is very important to know the lateral mixing in the layer.

Now, we will consider the variation of T-Cl relations at the sea surface. T-Cl relations at the sea surface on the meridian of $135^{\circ}40'E$ (G-line) and $137^{\circ}40'E$ (U-line) are

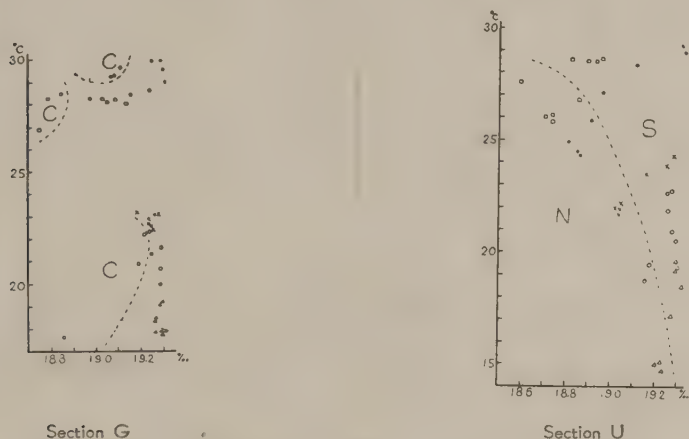


Fig. 17. T-Cl relations at the sea surface.

• August, 1956; × November, 1956; △ March, 1957; □ May, 1957.

C indicates the relation near the coast, that is north of $33^{\circ}N$.

N and S indicates the relation north and south of the current axis of the Kuroshio, respectively.

shown in Fig. 17. Along G-line, the difference between maximum and minimum water temperature is less than $2^{\circ}C$ and the difference in chlorinity is the largest in summer, while the difference in water temperature is nearly the same as that in summer and the difference in chlorinity is very small in winter. The transitional condition from winter to summer and from summer to winter is seen in spring and autumn, respectively. Along U-line, the same phenomena as those along G-line are found, but water temperature and chlorinity are clearly lower throughout the year in the upwelling region than in other region. Ichiye (1956) discussed the annual variation of chlorinity in the upper 50m layer in this sea area and concluded that the harmonic components of the annual variation were dependent upon lateral mixing and advective effect. Here, the decrease of chlorinity at the sea surface from winter to summer is treated on the basis of the diffusion equation, where the other effects except for lateral mixing are left out of consideration. If the decrease of chlorinity near the coast is presumed, it will be made clear how the effect extends off-shore. As the horizontal gradient of chlorinity at the sea surface is very small in winter, initial values are assumed to be independent on the distance from the coast. Further, the decrease of chlorinity at the

coast is assumed to be proportional to the time for simplification. Then we have

$$\frac{\partial S}{\partial t} = A \frac{\partial^2 S}{\partial y^2}$$

with

$$\begin{cases} S = kt & \text{at } y=0, \\ S=0 & \text{when } t=0, \end{cases}$$

where S , y and A is the amount of variation of chlorinity, the distance from the coast and the coefficient of horizontal diffusion, respectively. y is measured positively towards the offing. The solution of the above equation is (Carslaw & Jaeger, 1947)

$$\begin{aligned} S &= -kt \left[\left(1 + \frac{y^2}{2At} \right) \left\{ 1 - \Phi \left(\frac{y}{2\sqrt{At}} \right) \right\} - \frac{y}{\sqrt{\pi At}} e^{-\frac{y^2}{4At}} \right] \\ &= 4kt i^2 \operatorname{erfc} \frac{y}{2\sqrt{At}}. \end{aligned}$$

Here $i^2 \operatorname{erfc} x = \int_x^\infty i^{n-1} \operatorname{erfc} \xi d\xi$ and $\operatorname{erfc} \xi = 1 - \Phi(\xi)$.

Since $S_0 = kt$ at $y=0$, we have

$$\frac{S}{S_0} = 4i^2 \operatorname{erfc} \frac{y}{2\sqrt{At}} = 4i^2 \operatorname{erfc} \alpha,$$

where $\alpha = \frac{y}{2\sqrt{At}}$.

Thus, $\frac{S}{S_0}$ is indicated by a function of α and decreases as α increases. When α can be computed from $\frac{S}{S_0}$ which is obtained by repeated observations, A is also found since t and y are known.

The results should be applied to the variation along G-line from March 13~14 to August 1, 1957. When we choose the origin of y -axis at $33^\circ 20'N$, the values of α is obtained as in Table 4. The relation between α and y is shown in Fig. 18. If the assumption adopted above is satisfied strictly, a linear relation should be found between them because α is proportional to y . The result shown in Fig. 18 can be interpreted

Latitude	33°-20'	33°-00'	32°-40'	32°-20'	32°-00'	31°-40'	31°-20'	31°-00'N
S	-0.54	-0.45	-0.51	-0.22	-0.16	-0.30	-0.22	-0.24
S/S_0	1	0.83	0.94	0.41	0.30	0.56	0.41	0.44
α	0	0.07	0.02	0.37	0.48	0.24	0.37	0.34

Table. 4. The variations of surface chlorinity and the values of α along G-line from March to August, 1957.

as a linear one near the coast ($y < 80$ nautical miles), but the values of α in the offing ($y > 100$ nautical miles) are rather smaller than those at $y = 60$ and 80 nautical miles and a discontinuous variation is found between latter both places. So, it is understood that the decrease of chlorinity in the coastal region does not reach to the offing by lateral mixing and that the distance to which the effect extends is about 80 nautical miles from the coast in this case. This may connect with the oceanic conditions that the stream axis and the warm core is located at nearly $33^\circ N$ and $32^\circ N$ during this period, respec-

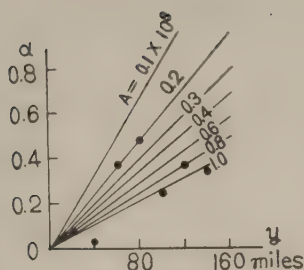


Fig. 18. Relation between α and y south of Shionomisaki from March to August, 1957.

on the distance from the coast, respectively. In this case, the rise of water temperature is given by

$$\Delta\theta_0 = \frac{2f_0}{K} \left(\frac{kt}{\pi} \right)^{\frac{1}{2}}$$

after section 3. Thus, the variation of T-Cl relation at the sea surface is found by using the variation of water temperature and chlorinity obtained above. Fig. 19

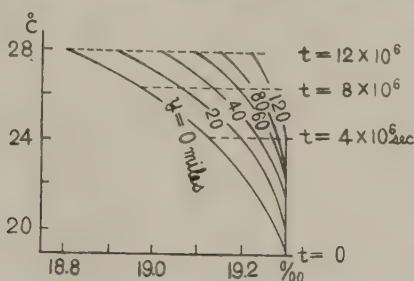


Fig. 19. Computed T-Cl relations at the sea surface when lateral mixing is considered. Here, $A = 0.2 \times 10^8$ is assumed.

shows the variation of T-Cl relation on the assumption as follows: initial values of water temperature and chlorinity are 19.0°C and 19.30‰ ; $S_0 = 0.50\text{‰}$, at $y = 0$; $A = 0.2 \times 10^8$; $f_0 = 4 \times 10^{-3}$; $k = 3$. Although the above results are calculated on the simple assumptions, the outline agrees nearly with the observed feature. So, more interesting results will be obtained if more reasonable assumptions are made.

6. Summary

Features in the distributions of water temperature and density in each season are indicated and differences of their seasonal variations between in the cold and warm region are made clear. The rise of water temperature near the surface is mainly due to the heat flux at the surface which is larger in the off-shore warm region than in the cold region. Water temperature plays an important role in the growth of homogeneous layer from summer to winter, but the influence of chlorinity upon the process is negligible. Furthermore, the fluctuations in T-Cl relations are treated and some interesting results are introduced.

Acknowledgement: In conclusion, the present author expresses his hearty thanks to

Prof. K. Hidaka, Dr. Y. Shibata, Dr. K. Hishida, Dr. T. Nan'niti and Dr. K. Yoshida for their kind advice and guidance.

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On the Forecasting of the Movement of the Typhoon

(3rd report)

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Abstract

For the purpose of computation of the general pressure gradient that plays an important role in the forecasting of the movement of the typhoon, we chose four representative points, which lie at a distance equal to the diameter of the circle 'S' (an inscribed circle of the outermost closed isobar) from the centre of a typhoon to north, south, east and west, respectively, and we take the pressure gradients at these four points as general pressure gradients and estimate the moving direction of the typhoon from the resultant of two moving directions on the surface and at the 500mb level. Also we found the relationship between the absolute value of the vector of the moving direction and the twice value of the diameter of the circle 'S'. And some trials were done applying the results to the practice of the forecasting for the interval of twenty-four hours.

Introduction

Previously, as a step to gain a method of forecasting of the movement of the typhoon, the author had made an attempt to separate the pressure field into the general pressure field $\bar{P}$ which has a large scale playing an important role in the movement of the typhoon and the small scale pressure field $P(n)$ which is accompanied with the typhoon, and have obtained a method to compute the general pressure field. Then, he investigated the relationship between the $\bar{P}$ and the velocity of progression of the typhoon. As the process of this method, however, is somewhat complicated, so it is not convenient for practical uses. Therefore, we tried to find a more simple and convenient method to the aim.

§ 1 The method of the computation of the moving direction.

By assuming the air is incompressible and frictionless, the equation of horizontal motion is transformed as follows:

$$\lambda S u = -\frac{1}{\rho} \frac{\partial}{\partial y} \iint_S p ds \quad (1)$$

$$\lambda S v = \frac{1}{\rho} \frac{\partial}{\partial x} \iint_S p ds \quad (2)$$

where λ : Coriolis parameter, S : area of vortex,
 ρ : density, x, u : coordinate, velocity toward east,
 y, v : coordinate, velocity toward north,

and $\bar{u} = \frac{1}{S} \iint_S u ds, \quad \bar{v} = \frac{1}{S} \iint_S v ds,$

then $\bar{u}$ and $\bar{v}$ are regarded as the velocity component of progression of the typhoon, if

it were considered as a votex.

As a approximate value of S we adopted ' S ' which is the maximum circle tangential to the closed isobar. Next, we chose four representative points, having a distance equal to the diameter of an inscribed circle of the outermost closed isobar such as to situate at north, east, south and west to the centre of the typhoon respectively. We take the pressure gradients at these four points as the approximate values of the general pressure gradients. At the 500mb chart we also adopt the same circle ' S ', where we take next equations instead of ①, ②

$$\lambda Su = -g \int \int Z ds \quad (3)$$

$$\lambda S \bar{v} = g \int \int Z ds \quad (4)$$

where Z : height of 500mb level.

The combination of the moving directions on the surface and at the 500mb is carried out as follows: From the origin of axes draw two lines parallel to the observed pressure gradient on the surface and on the 500mb level respectively, then the moving direction of the typhoon may be represented as the resultant direction of these two directions (observed pressure gradients are obtained from the pressure difference at four representative points, and in drawing the two lines the scale on the axes are thus the former by mm and the later by 100ft). And thus the moving direction of the typhoon is assumed to be obtained by the resultant of the two moving directions.

Next we show some practical examples, such as Typhoon No.14 and No. 15 in 1954 and some typhoons in the other years.

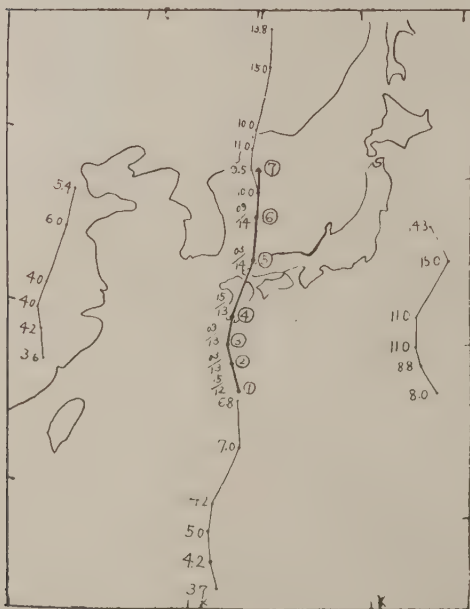


Fig. 1. Showing the path of Typhoon No. 12, September 12~14, 1954 and surface pressures at four representative points at each time

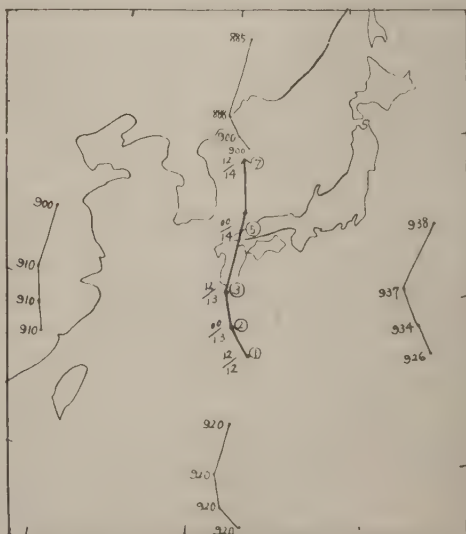


Fig. 2. Showing the path of Typhoon No. 12, September 12~14, 1954 and the heights of the 500mb level at four representative points at each time

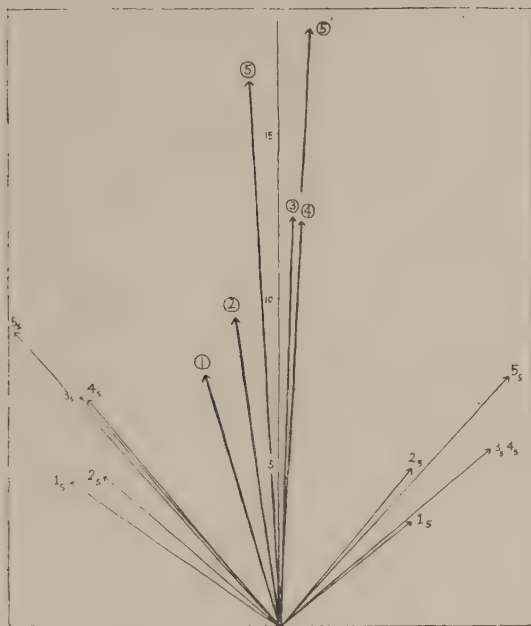


Fig. 3. Showing the moving direction of Typhoon No. 12 on the surface and at the 500mb level and the resultant direction of these two directions at each time. Suffix S and 5 denotes surface and 500mb level, respectively. (5)' denotes the predictional moving direction in Fig. 11

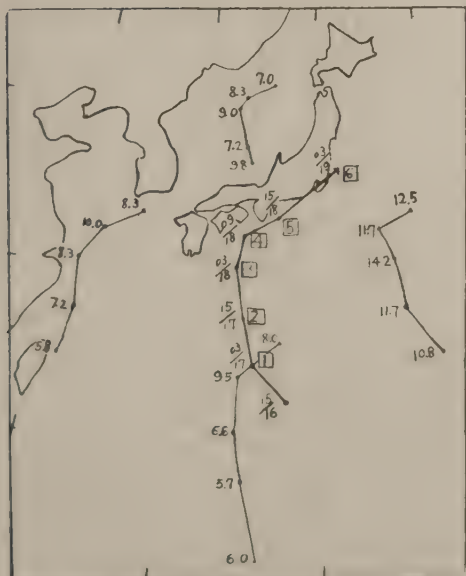


Fig. 4. Showing the path of Typhoon No. 14 September 16~19, 1954 and surface pressures at four representative points at each time

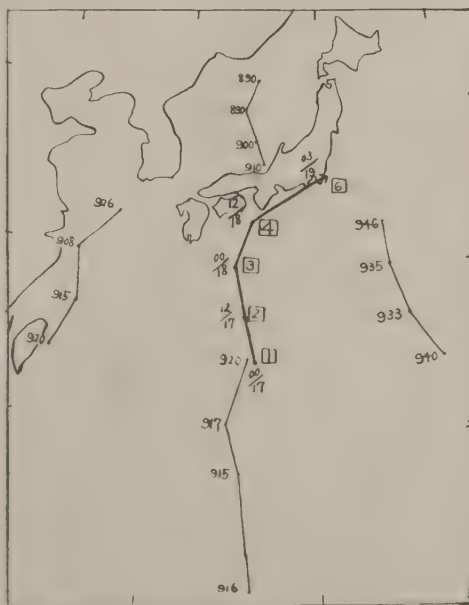


Fig. 5. Showing the path of Typhoon No. 14, September 17~19, 1954 and the heights of the 500mb level at four representative points at each time.

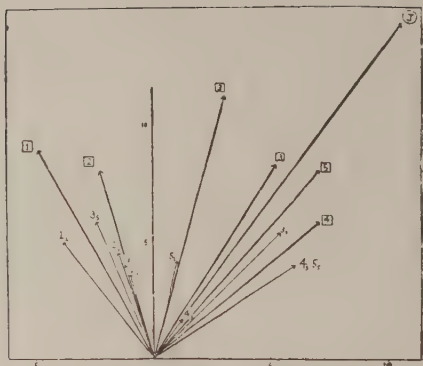


Fig. 6. Showing the moving direction of Typhoon No. 14 on the surface and at the 500mb level and the resultant direction of these two directions at each time. (J) denotes the moving direction of Typhoon Jane at 33.6N, 134.6E. (3)' denotes the predictional direction in Fig. 12.

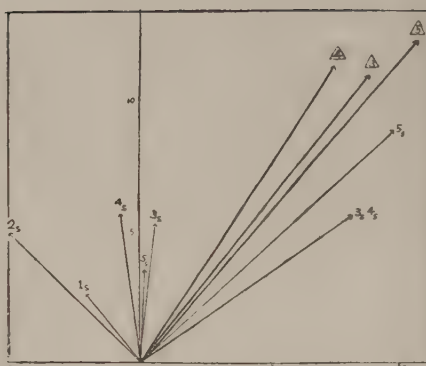


Fig. 9. Showing the moving direction of Typhoon No. 15 on the surface and at the 500mb level and the resultant direction of these two directions at each time

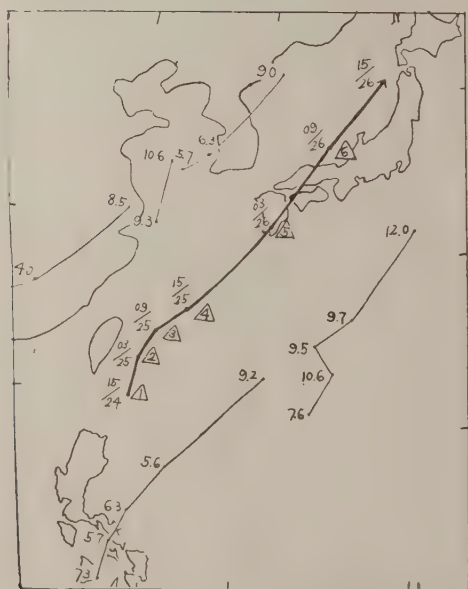


Fig. 7. Showing the path of Typhoon No. 15, September 24~26, 1954 and surface pressures at four representative points at each time.

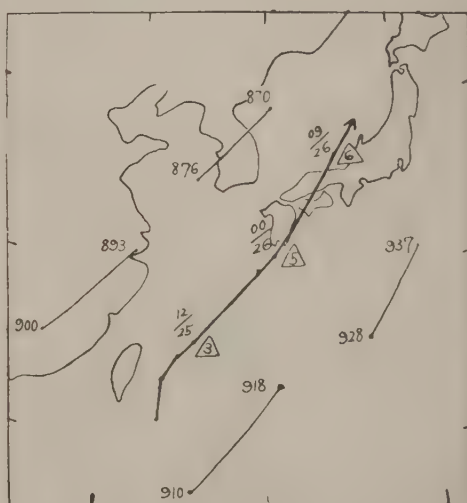


Fig. 8. Showing the path of Typhoon No. 15, September 24~26, 1954 and the heights of the 500mb level at four representative points at each time.

These figures show the actual path and the moving direction obtained by the above method in each typhoon.

The moving direction seems be fairly in accordance with actual direction from these figures. Especially, difficult points in forecasting described as below are well expressed in these figures: in the case of No. 12 it moved to northwards without recurving toward north-east after landing and in No. 14, it recurved toward north-east to east-north-east without moving to north wards as in Typhoon Jane.

§ 2 The method of the computation of the velocity

To estimate the velocity, we take the absolute value of the vector in the resultant moving direction on the axis of ordinate, and twice values of the diameter of the circle 'S' in a negative direction from a point on the positive axis of abscissa. On the right side of that point in the graph, we describe the observed moving distance of the typhoon for the interval of 12 hours (as a unit degree of longitude at 26°N was used). Then we draw a titling curve of equal velocity among these points. Inversely, we may obtain the velocity from this curve.

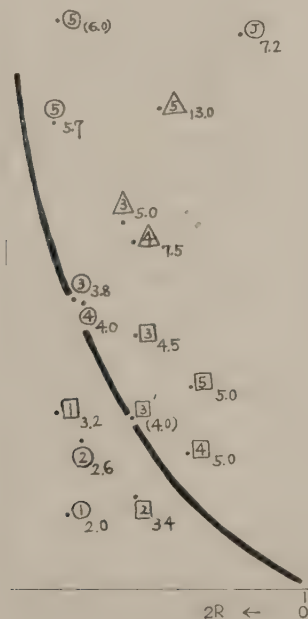


Fig. 10.

Showing the relationship between the absolute value of the vector of the resultant direction and twice diameter ($2R$) of the circle 'S' (the maximum circle tangential to the closed isobar. Suffix number shows the observed velocity of the typhoon for the interval of twelve hours in longitude (at 26°N). Full line shows the line of 4° in longitude approximately. (5)', (3)' denotes the predential velocity in Fig. 11 and Fig. 12, respectively.

Fig. 10 shows the velocities at each time in Typhoon No. 12, No. 14, No. 15 and Typhoon Jane.

From this figure we see that the velocity of Typhoon No. 12 was slow and Typhoon No. 14 did not move with fast velocity towards north as Typhoon Jane. But of Typhoon No. 15, the abnormal velocity is not well explained, perhaps it would be caused by the transformation into extratropical cyclone from the typhoon.

§ 3 On the successive forecasting

If we move the typhoon with the present velocity by a proper distance (according to the velocity and the state of the general pressure field) and presume pressures and heights at the four representative points in the new position of the typhoon., we might be able to estimate the velocity of the typhoon at the new position.

Repeating this method, we may forecast the moving of the typhoon during a fairly long period. Of course, a presumption of the values at the four representative points is fairly difficult, so it needs a future study (for instance, numerical forecasting)

But in the case of Typhoon No. 12 and No. 14, we may forecast if we take account of the curves of the values at the four points and the moving of the trough. Especially, it depends mostly upon the change of the height of the 500mb level whether the Typhoon recurves or not.

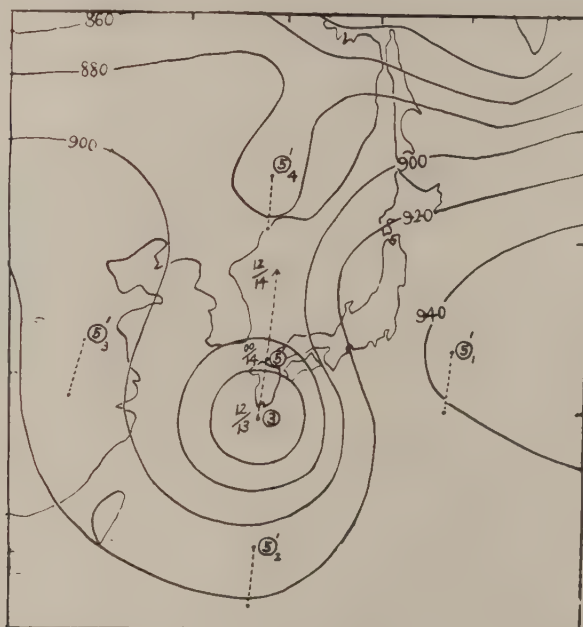


Fig. 11. The predisional moving direction of Typhoon No.12 and contours at the 500mb chart, 0000Z, September 13, 1954

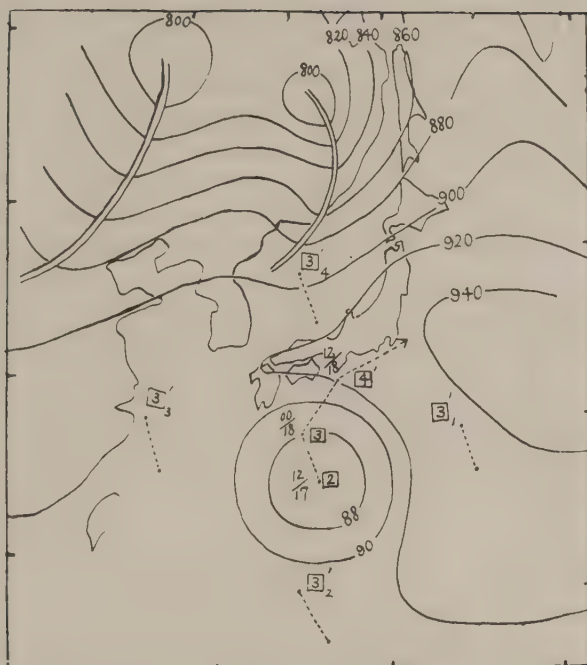


Fig. 12. The prediction moving direction of Typhoon No. 14 and contours at the 500mb chart, 0000Z, September 17, 1954.

Figures 11 and 12 show two practical examples which forecast the position of the typhoon after twenty-four hours.

The past, present and estimated future values of the height of the 500mb level and the surface pressures are also tabulated.

		example 1		500mb		
	12th 12h		13th 00h	13th 12h	14th 00h(estimated)	
⑤'₁	940(100 feet)		940	940	940	
⑤'₂	880		890	905	915	
⑤'₃	910		910	910	905	
⑤'₄	860		878	875	880	
		example 2		surface		
	12th 09h		12th 15h	13th 03h	13th 09h	14th 00h(estimated)
⑤'₁	1014.5(mb)		1013.5	1013.4	1016.2	1015.0
⑤'₂	850.0		990.0	994.0	999.0	1005.0
⑤'₃	1011.0		1008.0	1008.0	1007.0	1005.0
⑤'₄	1013.5		1013.5	1016.0	1014.0	1012.0
		example 3		500mb		
	15th 12h	16th 00h	16th 12h	17th 00h	17th 12h	18th 00h(estimated)
(3)'₁	940(100 feet)	936	940	941	936	938
(3)'₂	902	902	910	915	915	915

(3)' ₃	924	910	900	900	915	915
(3)' ₄	850	855	860	898	892	885

example 4 surface

	16th 09h	16th 15h	17th 03h	17th 09h	17th 15h	18th 00h(estimated)
(3)' ₁	1013.0(mb)	1010.0	1011.2	1013.0	1011.6	1012.6
(3)' ₂	999.0	996.0	999.0	1007.0	1005.0	1006.8
(3)' ₃	1012.0	1009.0	1011.0	1010.0	1009.0	1009.3
(3)' ₄	1001.0	1004.2	1007.0	1006.0	1008.8	1008.3

From the table, we notice that the presumption of the height of the 500mb level of (5)'₄ differs from that of (3)'₄. That is to say, the values in (5)'₄ is high, and this due to assumptions we took that in (5)'₄ a weak trough reached at 13th 12h and a ridge will reach later, while in (3)'₄ a ridge reached at 17th 00h and a trough will reach later gradually. Further, it may be expected that when the typhoon reaches (4)', the value in (4)' will become more lower. Accordingly, it is thought that the typhoon moves along the coast of Pacific as recurving stronger toward east.

Conclusion

We deem that the forecasting of the progression of the typhoon may be done with passable accuracy for the interval of 12 hours, but the forecasting for the longer interval claims still further study. Besides, it is important that we should estimate the values of the four points objectively.

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Kuroshio in Enshūnada

(2nd Report. On the T - σ_t relation)

SHUNJI KONAGA.

1. Introduction.

Recently, The oceanographic observations, using the bathythermograph have been carried out, and then we have many opportunities to observe the water temperature alone, from which it is preferable to know the density of sea water approximately. This problem was discussed in the sea south of Honshū by Yasui (1955). He formulated the chlorinity as a function of the water temperature, under the relation that T - Cl diagram had the same form in the same water mass. And then he showed σ_t as a function of the temperature alone, and computed the dynamic depth anomalies in a simplified form as follows:

$$\Delta D_{\infty} \int_0^D T dT$$

It is very useful, but his formulae are based on T - Cl relationship, which is so complicated that we try to describe σ_t , computed by means of the Kunudsen's table, as a function of the temperature alone in this paper.

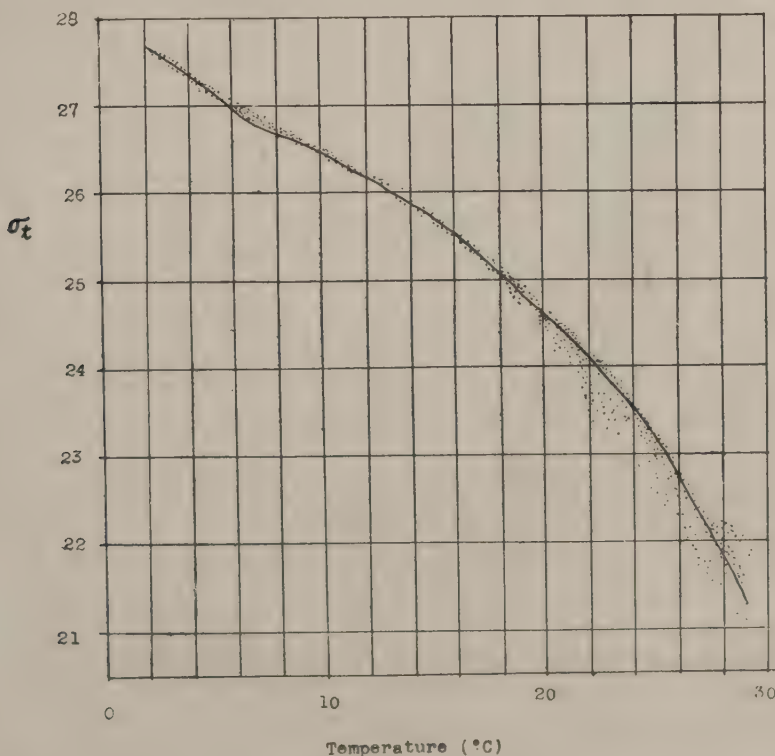


Fig. 1 T - σ_t diagram in the sea south of Honshū (Shumpū-maru, in May, 1954--November, 1956) .

2. Calculation of the density *in situ* from the temperature

In Fig. 1, we plot σ_t against the temperature in the sea south of Honshū, using the data on board Shumpu-maru in 1954–1956. It may be divided into three parts as follows:

$$(1) \quad T > 22^\circ\text{C}, \quad (2) \quad 22^\circ\text{C} > T > 8^\circ\text{C}, \quad (3) \quad 8^\circ\text{C} > T > 2^\circ\text{C}.$$

Then we obtain a fitting curves by means of the method of least squares, respectively, as shown in the following formulae:

$$\sigma_t = 20.85 + 0.568T - 0.0191T^2 \quad T > 22^\circ\text{C} \quad (1)$$

$$= 26.96 + 0.0167T - 0.0067T^2 \quad 22^\circ\text{C} > T > 8^\circ\text{C} \quad (2)$$

$$= 27.90 - 0.070T - 0.0229T^2 + 0.00156T^3 \quad 8^\circ\text{C} > T > 2^\circ\text{C} \quad (3)$$

The plotted values scatter little from the curves according to (1)–(3), excluding at and near the surface layer. In these layers, σ_t are unable to be shown definitely by the temperature alone in a simple form.

In table 1, the values of σ_t , calculated from the formulae (1)–(3), are shown in the first column and these from standard method in the second column in three hydrographic stations shown in Fig. 3.

Here A is located in the Kuroshio-current, B, in the south of Kuroshio-current and C, in the north of Kuroshio-current.

Next, we consider the error in σ_t due to the neglect of the chlorinity. From the Knudsen's formula:

$$\sigma_t = \Sigma_t + (\sigma_0 + 0.1324) [1 - A_t + (\sigma_0 - 0.1324)B_t] \quad (4)$$

$$\text{where } \Sigma_t = -\frac{(t-3.98)^2}{503.570} \cdot \frac{(t+283)}{(t+67.26)}$$

$$A_t = t(4.7867 - 0.098185t + 0.0010843t^2) \times 10^{-3}$$

$$B_t = t(18.030 - 0.8164t + 0.01667t^2) \times 10^{-6}$$

$$\text{and } \sigma_0 = -0.069 + 1.4708Cl - 0.01570Cl^2 + 0.0000398Cl^3$$

we obtain its derivatives, $\partial\sigma_t$, as temperature being constant, in the following:

$$\partial\sigma_t = \text{const} = \partial\sigma_0(1 - A_t) + 2\partial\sigma_0 \cdot B_t$$

Now in T-Cl diagram, we may consider that the chlorinity scatters in the range of 0.1‰ at the same temperature, that is, $\partial C = 0.05\%$ may be taken in this case.

For example, we consider the following case:

$$Cl = 19.10\text{‰}, \quad T = 0^\circ\text{C}, \quad 10^\circ\text{C} \quad \text{and} \quad 20^\circ\text{C}, \quad \text{respectively.}$$

$$\partial\sigma_0 \quad \text{is equal to } 0.07,$$

$$\partial\sigma_{10} \quad \text{is equal to } -0.068,$$

$$\text{and } \partial\sigma_{20} \quad \text{is equal to } 0.065, \quad \text{respectively.}$$

Consequently, we may consider that σ_t has the error at the most of 0.1 in our method.

3. Application to the computation of the dynamic depth anomalies.

The specific volume *in situ*, $\Delta\sigma_t$, is described as follows:

$$\Delta\sigma_t = 0.02736 - \frac{\sigma_t \times 10^{-3}}{1 + \sigma_t \times 10^{-3}} \quad (5)$$

Substituting from (1)–(3) into (5) and neglecting the higher orders, we obtain the following relations respectively,

$$J_{st} = 0.00695 - 0.000545T + 0.0000186T^2 - 0.000000022T^3 + 0.0000000036T^4 \quad T > 22^\circ C \quad (6)$$

$$= 0.00113 - 0.000017T + 0.0000064T^2 \quad 22^\circ C > T > 8^\circ C \quad (7)$$

$$= 0.0024 - 0.000066T + 0.0000227T^2 - 0.0000015T^3 \quad 8^\circ C > T > 2^\circ C \quad (8)$$

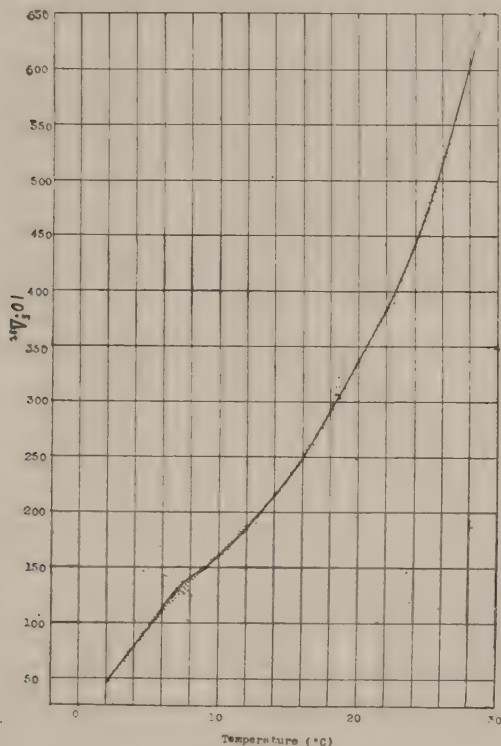


Fig. 2 T- J_{st} diagram in the sea south of Honshū.

In Fig. 2, the values of J_{st} , calculated from the above formulæ (in full line), and those from the standard method are shown. The values at the station A, B and C, are also shown in table 2.

The character in J_{st} seems to be similar to that of σ_t in these figures and tables.

And then, we calculate the depth anomalies, J_D , using above relations. The results are in good agreement with those from the standard method as shown in table 3. Computing the standard deviation of the differences between them in this table, we obtain 2 dynamic centimeters.

For example, we show the distribution of dynamic depth anomalies in August September, 1955, computed from the standard method in Fig. 3 (a) and those from our method (b).

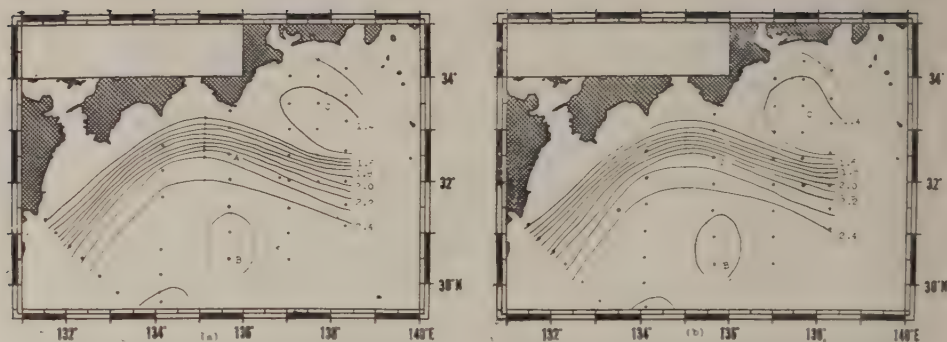


Fig 3 Dynamic depth anomalies obtained from (a) the standard method, (b) our method.

4. Acknwldgmnt

The author wishes to express his sincerly thanks to Dr. Hishida and Mr. Moriyasu for their kind encouragement to bring this paper.

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Table 1. The value of density *in situ*, calculated from the formulae (1), (2) and (8), and by the standard method in the sea south of Honshu, A is located in the Kuroshio current, B, south of the Kuroshio current, and C, north of the Kuroshio current in the cold water region in Enshu-nada.

A		B		C	
from (1)-(8)	Knudsen's table	from (1)-(8)	Knudsen's table	from (1)-(8)	Knudsen's table
21.55	21.60	21.90	21.92	22.70	22.23
21.55	21.55	21.90	21.92	22.70	22.25
21.55	21.58	22.15	22.06	24.50	24.23
22.00	21.98	23.50	23.41	25.25	25.15
22.45	22.39	24.10	24.09	25.50	25.50
23.30	23.25	24.30	24.37	25.75	25.71
24.20	24.14	24.55	24.61	26.10	26.09
24.75	24.67	24.70	24.77	26.40	26.38
25.55	25.56	25.00	25.00	26.70	26.72
26.00	25.98	25.30	25.32	26.95	26.92
26.50	26.50	25.65	25.68	27.10	27.08
26.85	26.86	26.10	26.08	27.20	27.20
27.15	27.16	26.75	26.73	27.35	27.33
27.35	27.35	27.10	27.10	27.45	27.43

Table 2. The value of specific volume anomalies *in situ*, calculated from (6)-(8), and those from the standard method at the station of A, B and C respectively.

A		B		C	
from (1)-(8)	stand. method	from (1)-(8)	stand. method	from (1)-(8)	stand. method
637	622	605	591	512	561
637	626	602	591	512	559
637	624	580	578	350	370
577	585	445	449	275	283
540	546	387	384	242	249
455	464	365	357	225	230
375	379	342	334	192	193
325	328	327	319	165	166
237	244	300	297	140	134
200	204	267	267	120	115
150	154	232	232	100	99
125	120	192	194	85	88
90	92	140	133	72	76
75	74	97	98	62	66

Table 3. The value of dynamic depth anomalies calculated from our method and those from standard method, in the sea south of Honshu in August-September, 1955.

our method	stand. method	our method	stand. method	our method	stand. method
2.346	2.338	2.239	2.245	1.839	1.861
2.443	2.433	2.388	2.379	1.355	1.366
2.440	2.431	2.488	2.492	1.214	1.235
2.411	2.400	2.571	2.574	2.125	2.136
2.425	2.402	2.544	2.527	2.463	2.501
2.372	2.350	2.494	2.511	2.235	2.246
2.379	2.346	2.392	2.432	1.702	1.733
2.437	2.401	2.476	2.422	1.371	1.422
2.436	2.441	2.233	2.266	1.346	1.390
2.489	2.415	1.545	1.524	1.417	1.428
2.220	2.240	1.385	1.417	1.610	1.647
2.184	2.147	1.310	1.336		
1.736	1.736	1.482	1.466		
1.514	1.493	1.473	1.502		
1.524	1.547	1.423	1.438		

Errata and addition

Page	line	for	read
Contents	3	Shigeru	Shigeo
1	13	chlorininty	chlorinity
1	16	ured	used
2	5	regin	region
2	9	sbove	above
2	12	transi tional	transitional
3	Fig. 5	1956	1957
5	2	dengity	density
5	3	tha	the
5	12	unstadle	unstable
5	*11	,too,	,too.
5	* 4	sectins	sections
6	* 3	gvien	given
7	5	fois	fo is
7	5	flax	flux
7	6	conductioity	conductivity
7	8	$\theta = -\frac{f_0}{K} \left[z \left\{ 1 - \phi \left(\frac{z}{2\sqrt{kt}} \right) \right\} - \frac{2\sqrt{kt}}{\sqrt{\pi}} e^{-\frac{z^2}{4kt}} \right]$	$\theta = -\frac{f_0}{K} \left[z \left\{ 1 - \phi \left(\frac{z}{2\sqrt{kt}} \right) \right\} - \frac{2\sqrt{kt}}{\sqrt{\pi}} e^{-\frac{z^2}{4kt}} \right]$
7	* 9	Φ_1 is the	Φ is the
7	* 4	$\frac{\Delta \theta}{\beta z_0} = \dots\dots$	$\frac{\Delta \theta z}{\beta z_0} = \dots\dots$
7	* 1	found then	found and then
8	Fig.11	$\alpha = \frac{z}{2\sqrt{kt}} = \alpha z$	$\alpha = \frac{z}{2\sqrt{kt}} = \alpha z$
9	13	previons	previous
9	*13	$\frac{\Delta \alpha_0}{\alpha_0} \div 0.05$	$\frac{\Delta \alpha_0}{\alpha_0} \div 0.05$
9	* 3	as computed	is computed
10	2	oberved	observed
10	* 5	$\frac{\partial \theta'}{\partial t} \Delta t =$	$\frac{\partial \theta'}{\partial t} t = t \Delta t$
11	*12	Fig. 12	Fig. 12
13	7	extenal	external
14	6	isttu	situ
15	Fig.17	□ May, 1957	○ May, 1957
16	9	$\frac{y}{\sqrt{\pi A t}} e^{-\frac{y^2}{4 A t}}$	$\frac{y}{\sqrt{\pi A t}} e^{-\frac{y^2}{4 A t}}$
18	* 1	Oceenagr.	Oceanogr.
20	2	approximate	approximate
20	8	insteacl	instead
20	18	mm	mb
24	5	akle	ab'le
27	9	oo	oc
28	6	squares	squares
28	*13	(1-At)	(1-At)
28	*11	$\partial C = 0.05\%$	$\partial CI = 0.05\%$

* from the bottom

